Arctic Climate Feedback Response to Local Sea-Ice Concentration and Remote Sea Surface Temperature Changes in PAMIP Simulations

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Abstract

Local and remote processes have been suggested to drive Arctic amplification (AA) – the enhanced warming of the Arctic region relative to other areas under increased greenhouse gases. We use Polar Amplification Model Intercomparison Project (PAMIP) simulations with changes in Arctic sea-ice with fixed global sea surface temperature (SST), or changes in global SST with fixed Arctic sea-ice to untangle the climate response to Arctic sea-ice loss or SST-induced warming, respectively. In response to Arctic sea-ice loss, the surface albedo feedback activates in summer mainly to increase oceanic heat uptake, leading to weak summertime warming. During winter, Arctic sea-ice loss greatly enhances oceanic heat release, which produces Arctic bottom-heavy warming and triggers positive lapse rate and cloud feedbacks, leading to large AA. In contrast, enhanced atmospheric energy convergence into the Arctic becomes the dominant contributor to relatively small AA under global SST-induced warming. Water vapor feedback contributes to Arctic warming but opposes AA due to larger tropical than Arctic moistening under SST-induced warming with fixed Arctic sea-ice. We also find top-heavy to uniform (bottom-heavy) Arctic warming and moistening in the Arctic mid-upper (lower) troposphere in the SST (Arctic sea-ice) perturbation runs, producing a negative-neutral (positive) Arctic lapse rate feedback, respectively. Lastly, we show that the responses to global SST or polar SIC perturbations are linearly separable. Our results suggest that large AA is caused primarily by sea-ice loss and resultant local changes in surface fluxes, while increased poleward energy transport can only produce weak AA under fixed sea ice.

Keywords: Arctic amplification; sea-ice loss; climate feedback; global warming; Arctic warming; ocean heat release; atmospheric energy transport

1. Introduction

The Arctic region warms faster than the rest of the world in response to increased greenhouse gas (GHG) concentrations – a phenomenon known as Arctic amplification (AA) (Serreze and Barry 2011; Walsh 2014; England et al. 2021; Taylor et al. 2022). Many mechanisms have been proposed to explain AA such as surface albedo feedback (Hall 2004; Winton 2006), enhanced poleward energy transport (Cai 2005; Henry et al. 2021), increased surface downwelling longwave (LW) radiation (Burt et al. 2016; Gong et al. 2017), Arctic positive lapse rate feedback (Pithan and Mauritsen 2014; Goosse et al. 2018), and sea-ice loss (Deser et al. 2010; Kumar et al. 2010; Screen and Simmonds 2010a, b; Boeke and Taylor 2018; Dai et al. 2019). The local and remote mechanisms suggested to contribute to AA are tightly coupled (Feldl et al. 2017b; Henry et al. 2021; Dai and Jenkins 2023), making the exact causes of AA unclear in a fully coupled system. For instance, sea-ice loss and the spatial patterns of surface warming largely shape Arctic positive lapse rate feedback (Feldl et al. 2020; Boeke et al 2021; Jenkins and Dai 2021). Further, warming in low-mid latitude regions influences Arctic mid-upper tropospheric warming through changes in atmospheric energy convergence into the Arctic, affecting the structure of Arctic warming profiles and lapse rate feedback (Perlwitz et al. 2015; Feldl et al. 2020; Hay et al. 2022). Thus, more work is needed to understand how local and remote processes influence Arctic warming and AA.

Arctic sea-ice loss plays an essential role in local Arctic warming (Dai et al. 2019; Linke et al. 2023b) and may contribute to warmer winters in northern hemisphere mid-latitude areas (Sun et al. 2016). Additionally, Arctic sea-ice loss may weaken the stratospheric polar vortex, which can affect weather patterns in the midlatitudes (Liang et al. 2023). As sea-ice retreats, increased energy transfer from warm, open water surfaces to the frigid overlying atmosphere during polar night contributes to large AA (Kumar et al. 2010; Deser et al. 2010; Screen and Simmonds 2010a, b; Boeke and Taylor 2018; Taylor et al. 2018; Dai et al. 2019; Dai and Jenkins 2023). Specifically, Dai et al. (2019) showed that AA weakens in model experiments with 1%/year CO_2 increases and fixed SIC for surface flux calculations, and that negligible additional AA will occur after sea-ice completely melts away. Davy and Griewank (2023) confirmed this finding by showing that as the rate of sea-ice loss decreases in the future, concurrent AA weakens.

Another process underlying AA is the lapse rate feedback that depends on local vertical warming structures (Pithan and Mauritsen 2014; Linke et al. 2023a; Zhou et al. 2023). Under a bottom-heavy warming profile, outgoing LW radiation at the top of the atmosphere (TOA) is reduced relative to vertically uniform warming, thereby enhancing surface warming (Boeke et al. 2021; Dai and Jenkins 2023). In contrast, a top-heavy warming profile, as seen in the tropics, suppresses surface warming by increasing outgoing LW radiation (Colman and Soden 2021). The lapse rate feedback has been considered as a major contributor to AA due to its large Arctic versus tropical warming effect (Pithan and Mauritsen 2014; Goosse et al. 2018; Hahn et al. 2021). Previous studies have attributed Arctic bottom-heavy warming and the resultant positive lapse rate feedback to high lower-tropospheric stability, which effectively traps warming at the surface (Bintanja et al. 2011; Pithan and Mauritsen 2014). However, recent studies suggest that Arctic lapse rate feedback is strongly correlated with surface warming patterns and sea-ice loss (Feldl et al. 2020; Boeke et al. 2021; Jenkins and Dai 2021) rather than stability strength (Jenkins and Dai 2022; Dai and Jenkins 2023). Remote processes, such as enhanced moist static energy convergence into the Arctic, may also influence Arctic lapse rate feedback by favoring warming in the mid-upper troposphere (Feldl et al. 2020), leading to negative lapse rate feedback.

During summer, surface albedo and water vapor feedbacks activate in the Arctic in response to greenhouse gas (GHG) forcing. The surface albedo feedback makes a large positive contribution to Arctic energy imbalance in summer (Hall 2004; Winton 2006; Pithan and Mauritsen 2014; Goosse et al. 2018; Hahn et al. 2021); however, most of the enhanced shortwave (SW) absorption preferably warms the ocean mixed layer rather than near-surface air (Dai 2021; Dai and Jenkins 2023). Additionally, water vapor feedback has been suggested to contribute to Arctic *warming* (Ghatak and Miller 2013; Gong et al. 2017) but oppose Arctic *amplification* due to larger moistening in tropical regions than polar areas under increased GHGs (Pithan and Mauritsen 2014; Hahn et al. 2021). Jenkins and Dai (2022) showed that water vapor feedback and sea-ice loss spatial patterns are weakly correlated in ERA5 reanalysis data, but they did not quantify the underlying local and remote drivers of Arctic water vapor feedback. An improved understanding of Arctic water vapor feedback is needed as it enhances Arctic surface warming and melts sea ice, indirectly contributing to AA through the sea-ice feedback (Dai et al. 2019; Dai and Jenkins 2023). Moreover, water vapor feedback may interact with other processes by changing patterns of atmospheric latent energy transport (Chung and Feldl 2023) or amplifying other climate feedbacks (Beer and Eisenman 2022).

Cloud feedback impacts TOA and surface energy fluxes (Wetherald and Manabe 1988), but their response to local and remote processes is not fully understood. Previous studies have found an increase in local Arctic low cloud amounts and cloud water content in response to local sea-ice loss due to strong cold season ocean-atmosphere coupling (Schweiger et al. 2008; Kay and Gettelman 2009; Eastman and Warren 2010; Liu et al. 2012; Taylor et al. 2015; Kay et al. 2016; Morrison et al. 2018, 2019; Jenkins and Dai 2022; Jenkins et al. 2023; Taylor and Monroe 2023). Increased surface downwelling LW radiation from local Arctic cloud increases slows sea ice growth during Arctic autumn and winter, lengthening exposure of open water surfaces to heat the overlying air during the cold season (Monroe et al. 2021). Nonlocal cloud feedbacks may also contribute to Arctic warming and AA by affecting remote surface warming patterns and thus atmospheric energy transport into the Arctic (Vavrus et al. 2004; Middlemas et al. 2020).

Increased energy transport from midlatitudes into the Arctic has been suggested to influence AA (Cai 2005; Roe et al. 2015; Feldl et al. 2017b; Soldatenko 2021). Without sea-ice loss and associated surface heating, enhanced poleward atmospheric energy transport produces only weak AA in model simulations (Alexeev et al. 2005; Merlis and Henry 2018; Henry et al. 2021). On the other hand, inclusion of sea-ice loss effects in model simulations reduces atmospheric energy transport into the Arctic due to decreased temperature gradients between middle and high latitudes (Hwang et al. 2011; Jenkins and Dai 2021; Hahn et al. 2023). Further, Cardinale and Rose (2023) showed that an increase in the fraction of the Arctic energy convergence used to heat the surface may overcome the total decrease in Arctic energy convergence, contributing to winter Arctic warming. Inhomogeneous spatial patterns of radiative forcing also influence atmospheric poleward energy transport (Stuecker et al. 2018; Virgin and Smith 2019). When radiative forcing is negative in the Arctic, atmospheric poleward energy transport increases to offset the energy imbalance, inducing small AA (Virgin and Smith 2019). Additionally, Stuecker et al. (2018) found that atmospheric energy transport became an important contributor to AA in response to radiative forcing applied only in midlatitudes in fully coupled simulations, but they did not examine the effects of sea-ice loss in shaping the Arctic warming in response to such forcing.

The relative importance of sea-ice loss, positive climate feedbacks, and atmospheric energy transport in shaping AA is still debated and merits further investigation. Arctic climate feedbacks have been estimated in coupled model simulations (Pithan and Mauritsen 2014; Goosse et al. 2018; Stuecker et al. 2018; Previdi et al. 2020; Hahn et al. 2021); however, the influence of local sea-ice loss or remote SST warming on climate feedbacks cannot be explicitly quantified in a fully coupled system. To address these points, we use atmosphere-only simulations from the Polar Amplification Model Intercomparison Project (PAMIP; Smith et al. 2019) to answer the following questions:

1. What are the impacts of local Arctic SIC changes through enhanced oceanic heating of the atmosphere or global SST changes and background warming in atmosphere-only model simulations on Arctic surface warming, AA, radiative climate feedbacks, and atmospheric energy transport?
2. Do the individual responses to SST warming or Arctic SIC loss sum to the total response to the combined influences of SST warming and Arctic SIC loss occurring simultaneously?

The PAMIP experiments allow us to separate the climate response to perturbations in local sea ice or remote SST changes in model simulations under fixed GHG concentrations. The SST perturbation runs represent the climatic effects of background global warming without large AA, while the Arctic SIC change simulations show the impact from Arctic sea-ice loss without background global warming.

2. Methods

2.1 PAMIP experiments

We investigate how changes in global SST and/or local SIC impact Arctic surface warming, AA, climate feedbacks, and atmospheric energy transport using PAMIP atmosphere-only time slice experiments (Table 1; Smith et al. 2019). PAMIP experiment 1.1 (pdSST-pdSIC) serves as the control run where global SST and polar (i.e., Arctic and Antarctic) SIC fields are fixed at their present-day (pd) (i.e., year 2000) values. To isolate the response to global SST changes, we compare the pdSST-pdSIC run to PAMIP experiments 1.3 (piSST-pdSIC) and 1.4 (futSST-pdSIC) where polar (i.e., Arctic and Antarctic) SIC remains fixed at present-day conditions and SSTs over open water surfaces are set to preindustrial (pi) and future (fut) states (defined below), respectively. Likewise, we difference the pdSST-pdSIC run with PAMIP experiments 1.5 (pdSST-piArcSIC) and 1.6 (pdSST-futArcSIC) where SSTs outside the Arctic region are fixed at their present-day values and Arctic SIC is changed to preindustrial and future states to separate the impacts of sea-ice loss from other forcings. For the pdSST-piArcSIC and pdSST-futArcSIC simulations, SSTs are specified at their preindustrial or future values in regions where preindustrial or future SIC deviates by more than 10% of the present-day state, respectively (Smith et al. 2019).

Figure 1 shows the maps of prescribed SST and SIC changes for the preindustrial (Fig. 1a, b) or future (Fig. 1c, d) cases. To facilitate comparison with the future changes, which are relative to present-day, the historical changes are computed as present-day minus preindustrial in Fig. 1 and all other figures. We also compute the difference between pdSST-pdSIC and experiment 1.2 (piSST-piSIC; referred to as TOTAL) where global SSTs and polar SIC are changed simultaneously to their preindustrial states. We compare the results from the piSST-piSIC run to the sum of piSST-pdSIC, pdSST-piArcSIC and pdSST-piAntSIC (referred to as SUM) simulations to assess the linearity of the total climate response to both polar SIC and global SST changes. The preindustrial, present-day, and future time periods correspond to estimated Arctic SIC and/or global SST conditions under global-mean surface temperatures of 13.67°C, 14.24°C, and 15.67°C, respectively (Smith et al. 2019), which correspond to a historical warming of 0.57^{o}C and a future warming of 1.43^{o}C relative to present-day. Their corresponding SIC changes are also much larger for the future case than the historical case (Fig. 1).

Table 1. Summary of PAMIP experiments used in the analysis (from Smith et al. 2019).

Model Simulation	Full Name	Description
1.1 pdSST-pdSIC	Present day sea surface temperature Present-day sea-ice concentration	Year 2000 global SST and polar SIC; control run.
1.2 piSST-piSIC	Preindustrial sea surface temperature Preindustrial sea-ice concentration	Historical global SST and polar SIC; assesses total climate response to SST and SIC changes.
1.3 piSST-pdSIC	Preindustrial sea surface temperature Present-day sea-ice concentration	Historical (1.3) and future (1.4) global SST with polar SIC fixed at year 2000 conditions; assesses role of background warming without sea-ice feedback.
1.4 futSST-pdSIC	Future sea surface temperature Present-day sea-ice concentration	

1.5 pdSST-piArcSIC	Present-day sea surface temperature Preindustrial sea-ice concentration	Historical (1.5) and future (1.6) Arctic SIC with global SST fixed at year 2000 conditions; assesses role of Arctic sea-ice feedback without background warming.
1.6 pdSST-futArcSIC	Present-day sea surface temperature Future sea ice concentration	
1.7 pdSST-piAntSIC	Present-day SST Preindustrial Antarctic SIC	Historical Antarctic SIC with global SST fixed at year 2000 conditions; assesses role of Antarctic sea-ice feedback without background warming.

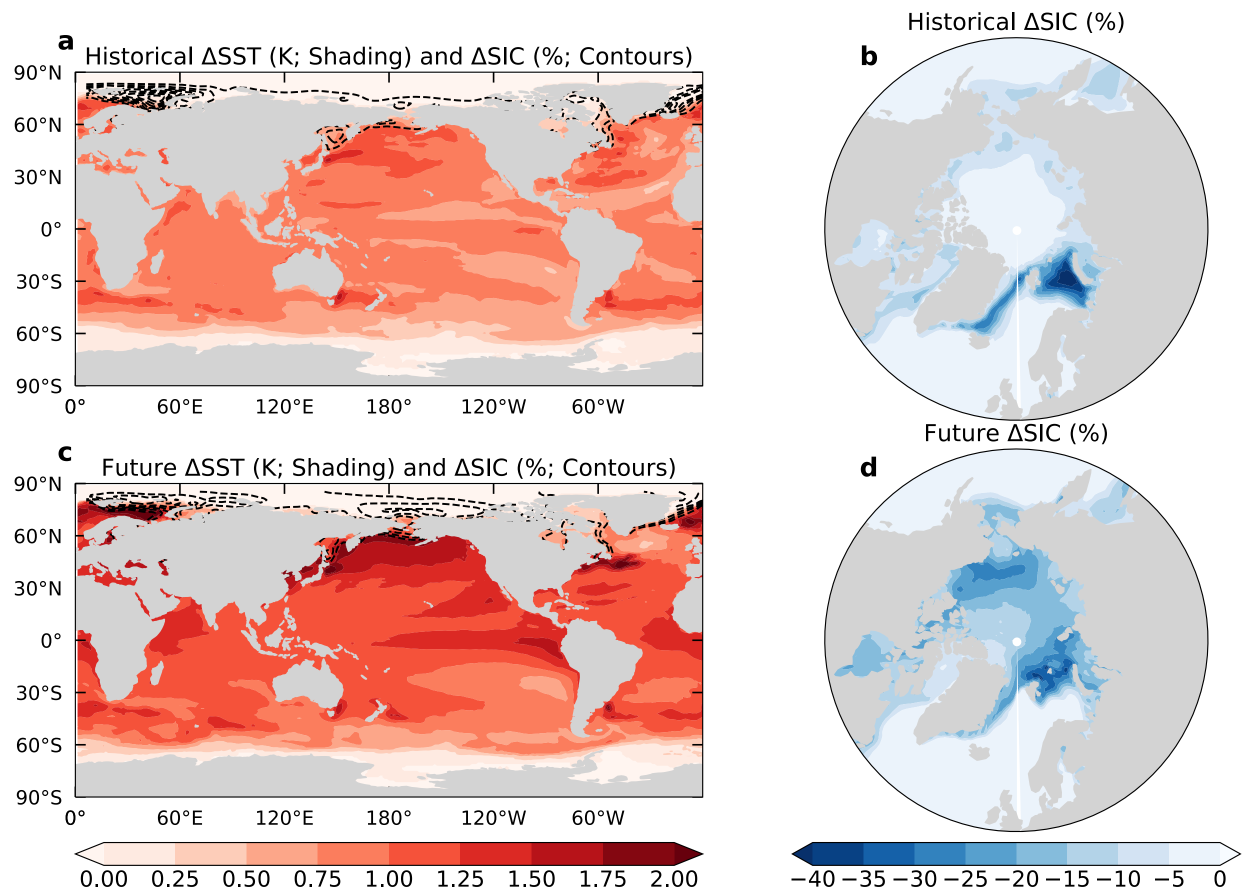

Fig. 1. (**a**, **c**) Annual mean changes in SST (K; shading) and Arctic SIC (%; contours; interval 5%) for the (**a**) historical (present-day minus preindustrial) and (**c**) future warming (future minus present-day) cases. Changes in SIC for the (**b**) historical and (**d**) future cases are shown as shading in (**b**) and (**d**) for clarity.

We use monthly-mean output from five models (i.e., AWI-CM1-1-MR, CESM2, CNRM-CM6-1, CanESM5, IPSL-CM6A-LR) that provided the necessary fields for our analysis. AWI-CM1-1-MR and CNRM-CM6-1 did not output the necessary variables for some calculations in piSST-piSIC (i.e., TOTAL) and is excluded in our comparison of piSST-piSIC to the sum of piSST-pdSIC, pdSST-piArcSIC, and pdSST-piAntSIC (i.e., SUM). Each model and experiment are initialized on 1 April 2000 and are run for 14-months, discarding the first two months as spin-up (Smith et al. 2019). To improve robustness of the results, we analyze the ensemble mean of the 100 ensemble runs with varied initial conditions for each model and experiment as atmospheric internal variability can mask the climatic response to SIC or SST changes (Screen et al. 2014). We define the Arctic region as the area poleward of 67°N following previous work (e.g., Dai et al. 2019; Jenkins and Dai 2022) because most Arctic sea-ice exists poleward of this latitude and the Arctic is mostly ocean surface in this region. We exclude land surfaces in our Arctic regional averages because surface warming is strongest over oceanic areas (Boeke and Taylor 2018; Dai et al. 2019) but inclusion of land areas does not qualitatively affect our results. Globally averaged fields include both land and ocean surfaces. For this study, we calculate AA as the difference between Arctic (excluding land) and global surface air temperature (ΔT_{as}) changes ($AA = \Delta T_{as,ARCTIC} - \Delta T_{as,GLOBAL}$) rather than as the ratio of Arctic to global warming to avoid dividing by near-zero values for global-mean surface air temperature changes.

2.2 Energy budgets

The vertically integrated energy budget equation (Eq. 1) for an atmospheric column accounts for the net TOA radiative flux ($R^{\downarrow}_{TOA}$; positive downward), net surface energy flux ($R^{\downarrow}_{SFC}$; positive downward), change in local energy storage in the atmospheric column ($\frac{\partial E}{\partial t}$), and horizontal convergence of energy ($-\nabla \cdot \boldsymbol{F}_A$) (Trenberth 1997; Fasullo and Trenberth 2008):

$$\frac{\partial E}{\partial t} = R^{\downarrow}_{TOA} - R^{\downarrow}_{SFC} - \nabla \cdot \boldsymbol{F}_A \quad , \tag{1}$$

where

$$E = \frac{1}{g}\int_{p_{TOA}}^{p_s} (c_p T + Lq + gz)\ dp. \tag{2}$$

In Eq. (2), E is the vertically integrated moist static energy, where c_pT, Lq, and gz denote atmospheric internal energy, latent energy, and potential energy, respectively. Atmospheric kinetic energy storage is small and is not included in Eq. (2), following previous studies (Oort and Vonder Haar 1976; Trenberth and Solomon 1994). For the flux terms, we calculate $R_{TOA}^{\downarrow}$ and $R_{SFC}^{\downarrow}$ as:

$$R_{TOA}^{\downarrow} = ASR^{\downarrow} - OLR^{\uparrow} \tag{3}$$

$$R_{SFC}^{\downarrow} = SW_{NET,SFC}^{\downarrow} - LW_{NET,SFC}^{\uparrow} - SH^{\uparrow} - LH^{\uparrow} \tag{4}$$

where $ASR^{\downarrow}$, $OLR^{\uparrow}$, $SW_{NET,SFC}^{\downarrow}$, $LW_{NET,SFC}^{\uparrow}$, $SH^{\uparrow}$, and $LH^{\uparrow}$ are the TOA absorbed SW radiation (positive downward), TOA outgoing LW radiation (positive upward), net surface SW radiation (positive downward), net surface LW radiation (positive upward), surface sensible and latent heat flux (positive upward), respectively. To estimate oceanic heat uptake (OHU), we calculate the net surface energy flux (Eq. 4) over ocean surfaces only.

We compute the horizontal atmospheric energy convergence ($-\nabla \cdot \boldsymbol{F}_A$) by rearranging the terms in Eq. (1) to obtain:

$$-\nabla \cdot \boldsymbol{F_A} = R_{SFC}^{\downarrow} - R_{TOA}^{\downarrow} + \frac{\partial E}{\partial t}. \tag{5}$$

Eq. (5) shows that the net convergence of the horizontal energy flux (in W m^{-2}) into a column is linked to the difference between the energy absorbed at the surface and net TOA radiation, and changes in local energy storage. We also calculate the atmospheric energy transport (AET; in PW) into the region north of a given latitude (ϕ) by taking the area integral of the net energy convergence over the region following previous studies (Hwang and Frierson 2010; Feldl et al. 2017a):

$$AET(\phi) = \int_{\phi}^{\pi/2} \int_{0}^{2\pi} \left(R_{SFC}^{\downarrow} - R_{TOA}^{\downarrow} + \frac{\partial E}{\partial t} \right) a^2 \cos\phi \, d\gamma d\phi. \tag{6}$$

In Eq. (6), a is the radius of Earth (~6.371×10^6 m), γ is the longitude, and ϕ is the latitude. *AET*(ϕ) represents the total energy crosses the latitude circle at ϕ (positive northward). For our Arctic region, ϕ=67°N.

2.3 Climate feedback calculations

The response of the atmospheric energy budget to a climate perturbation, assuming negligible changes in atmospheric energy storage, is:

$$\Delta R_{TOA}^{\downarrow} - \Delta R_{SFC}^{\downarrow} - \Delta(\nabla \cdot \boldsymbol{F_A}) = 0 \tag{7}$$

where $\Delta R^{\downarrow}_{TOA}$, $\Delta R^{\downarrow}_{SFC}$, and $\Delta(\nabla \cdot \boldsymbol{F_A})$ are changes in the net TOA radiative flux, net surface energy flux, and atmospheric horizontal energy convergence at each grid point, respectively (Stuecker et al. 2018; Hahn et al. 2021; Zhou et al. 2023). We use the Pendergrass et al. (2018) CESM1-CAM5 radiative kernels to decompose changes in the TOA net radiative flux into individual contributions from changes in surface albedo (ΔR_α), air temperature (ΔR_T), water vapor (ΔR_q), and clouds (ΔR_C):

$$\Delta R^{\downarrow}_{TOA} = \Delta R^{\downarrow}_{\alpha} + \Delta R^{\downarrow}_{q} + \Delta R^{\downarrow}_{T} + \Delta R^{\downarrow}_{C}. \tag{8}$$

GHG concentrations remain fixed at year 2000 levels in the PAMIP simulations, so we exclude an effective radiative forcing term from our TOA flux change decomposition. We also normalize the TOA flux changes in Eq. (8) by the local surface air temperature change (ΔT_{as}) to calculate the climate feedback parameter (λ_i) for each variable using:

$$\sum_i \lambda_i = \lambda_\alpha + \lambda_q + \lambda_T + \lambda_C = \frac{\Delta R^{\downarrow}_{\alpha} + \Delta R^{\downarrow}_{q} + \Delta R^{\downarrow}_{T} + \Delta R^{\downarrow}_{C}}{\Delta T_{as}} \tag{9}$$

For clarity, we use the term *feedback* to refer to the unnormalized TOA radiative flux changes (units: W m^{-2}) in Eq. (8) and *feedback parameter* to refer to the normalized TOA radiative fluxes (units: W m^{-2} K^{-1}) in Eq. (9).

Radiative kernels are computed by perturbing one climate variable in a radiative transfer model and keeping all other variables fixed to produce a TOA radiative flux response, which is divided by the amount of the perturbed variable change to derive the TOA flux change per unit variable change (Soden et al. 2008). To calculate the surface albedo feedback, we compute the product of the surface albedo kernel (K_α) and changes in surface albedo ($\Delta\alpha$): $\Delta R_\alpha = K_\alpha * \Delta\alpha$. For water vapor (Eq. 10) and temperature (Eq. 11) feedbacks, we vertically integrate the product of the kernel and change in each respective variable from the surface (p_s) to the tropopause (p_{TOA}):

$$\Delta R_q = \int_{p_{TOA}}^{p_s} K_q * \Delta \ln(q)\, dp \tag{10}$$

$$\Delta R_T = \int_{p_{TOA}}^{p_s} K_{T_a} * \Delta T_a\, dp \tag{11}$$

where q and T_a represent specific humidity and air temperature, respectively. Radiative emissions from water vapor scale with the natural logarithm of specific humidity, so we use $\Delta\ln(q)$ in Eq. (11) as done previously (Shell et al. 2008). Further, we assume that the tropopause pressure increases with latitude from 100 hPa at the equator to 300 hPa at the poles following Pithan and

Mauritsen (2014) to mask out the stratosphere. To calculate Planck and lapse rate feedbacks, we separate the temperature feedback (ΔR_T) into a component associated with vertically uniform warming equal to that of the surface (Planck feedback; ΔR_{PL}) and deviations from the vertically uniform warming profile (lapse rate feedback; ΔR_{LR}):

$$\Delta R_T = \Delta R_{LR} + \Delta R_{PL} = \int_{p_{TOA}}^{p_s} K_{T_a} * (\Delta T_a - \Delta T_{as})\, dp + \int_{p_{TOA}}^{p_s} K_{T_a} * \Delta T_{as}\, dp \qquad (12)$$

More details on Planck and lapse rate feedback calculations are provided in Dai and Jenkins (2023).

The change in cloud radiative forcing (ΔCRF) – the difference between all-sky and clear-sky radiative fluxes – provides a simple estimate of the energetic effects of clouds but does not represent cloud feedback as other processes also affect this difference (Soden et al. 2008; Block and Mauritsen 2013). To compute cloud feedback (ΔR_C), we subtract a cloud masking (CM) term from the ΔCRF to account for the effects of changes in surface albedo, temperature, and water vapor on ΔCRF (Soden et al. 2008):

$$\Delta R_C = \Delta CRF - CM \qquad (13)$$

where

$$CM = (K_\alpha - K_\alpha^C) * \Delta\alpha + \int_{p_{TOA}}^{p_0} \left(K_{T_a} - K_{T_a}^C\right) * \Delta T_a\, dp + \int_{p_{TOA}}^{p_0} \left(K_q - K_q^C\right) * \Delta l\, n(q)\, dp. \quad (14)$$

In Eq. (14) K_i and K_i^C are the all-sky and clear-sky kernels for surface albedo (α), air temperature (T_a), and water vapor (q). GHG concentrations are fixed in the PAMIP runs so we exclude a GHG masking term in Eq. (14).

2.4 Potential warming contribution estimates

To facilitate comparison, we quantify climate feedbacks, oceanic heat uptake, and horizontal atmospheric energy convergence in terms of their *potential* warming contributions following previous studies (e.g., Pithan and Mauritsen 2014; Goosse et al. 2021; Stuecker et al. 2018; Hahn et al. 2021). The potential warming contribution from the *i*th climate feedback ($\Delta T_i = \Delta R_i / \overline{\lambda}_{PL}$, in K) represents a hypothetic warming amount needed to rebalance the TOA energy flux change ($\Delta R_i = \lambda_i \Delta T_{as}$) through the negative Planck feedback at a new *equilibrium* state. Similarly, we can scale the other flux changes to estimate their potential warming contributions, and the total potential warming amount (ΔT) is estimated as (Goosse et al. 2018; Hahn et al. 2021):

$$\Delta T = -\frac{\sum_i \lambda_i \Delta T_{as}}{\bar{\lambda}_{PL}} - \frac{\lambda'_{PL}\Delta T_{as}}{\bar{\lambda}_{PL}} - \frac{\Delta(-\nabla\cdot \boldsymbol{F_A})}{\bar{\lambda}_{PL}} - \frac{\Delta OHU}{\bar{\lambda}_{PL}} \quad (15)$$

where $\bar{\lambda}_{PL}$ (in W m^{-2} K^{-1}) is the global-mean Planck feedback parameter and λ'_{PL} is the deviation of the local (λ_{PL}) Planck feedback parameter from its global mean: $\lambda'_{PL} = \lambda_{PL} - \bar{\lambda}_{PL}$. As noted by Dai and Jenkins (2023), this estimated warming amount often does not represent a real warming contribution as the TOA flux change (ΔR_i) may not be used to directly raise surface air temperature or the temperature response may be delayed. We average the terms in Eq. (15) over the Arctic (67°-90°N) and the tropics (23.5°S-23.5°N) to estimate the potential warming contribution of each process to surface warming and AA as done previously (Pithan and Mauritsen 2014; Goosse et al. 2018; Stuecker et al. 2018; Hahn et al. 2021).

3. Results

3.1 Surface warming response to changes in global SST or Arctic SIC

We first examine the annual-mean surface air temperature response to historical and future global SST (Fig. 2a, b) or Arctic SIC (Fig. 2c, d) changes shown in Figure 1. The globe experiences relatively uniform warming in pdSST-pdSIC relative to piSST-pdSIC (Fig. 2a, referred to as historical warming) and in futSST-pdSIC relative to pdSST-pdSIC (Fig. 2b, referred to as future warming), with slightly greater magnitude in the future SST case (around 1.0-2.0 K) than the historical case (0.5-1.0 K). Thus, the SST perturbation runs show background global warming without noticeable AA. In contrast, reduced Arctic sea-ice leads to large warming over the Arctic with little temperature change south of ~60°N in both the historical and future perturbed SIC runs (Fig. 2b, d). Note that the local Arctic warming is larger for the future case (~3-5 K; Fig. 2d) than the historical case (~1-3 K; Fig. 2c) as the future sea-ice loss is larger (Fig. 1c-d) and that the largest historical warming (Fig. 2c) occurs over the Barents-Kara Seas region where there is large sea-ice loss (Fig. 1b).

The seasonal cycle of the surface air temperature changes averaged over the Arctic (Fig. 3a) and globe (Fig. 3b) shows different responses to global SST or Arctic SIC perturbations. Global SST perturbations produce small Arctic warming during historical (~0.5-1.0 K) and future (~1.0-

2.0 K) periods for October-March and negligible summer warming (Fig. 3a). The global-mean surface temperature warms by ~0.8 K for the historical and ~1.2 K for the future SST cases, with little seasonal variation (Fig. 3b). Thus, there is small AA (<0.8 K) during October-March while the summer Arctic warming is weaker than the global-mean warming in the SST perturbation experiments (Fig. 3c). In contrast, Arctic sea-ice loss produces large Arctic warming from October-January for the historical (~3 K) and future (~6-8 K) cases, with weak warming in summer (Fig. 3a). Note that the peak warming shifts from October in the historical case to November in the future case. The global-mean warming response to the SIC changes is weak throughout most of the year except the cold season (Fig. 3b), which is due to the large warming in the Arctic (Fig. 2c-d). As a result, AA is strong from October-January for the two perturbed SIC cases, especially for the future SIC case (up to 7 K), while the AA is weak (<1 K) during the summer months (Fig. 3c).

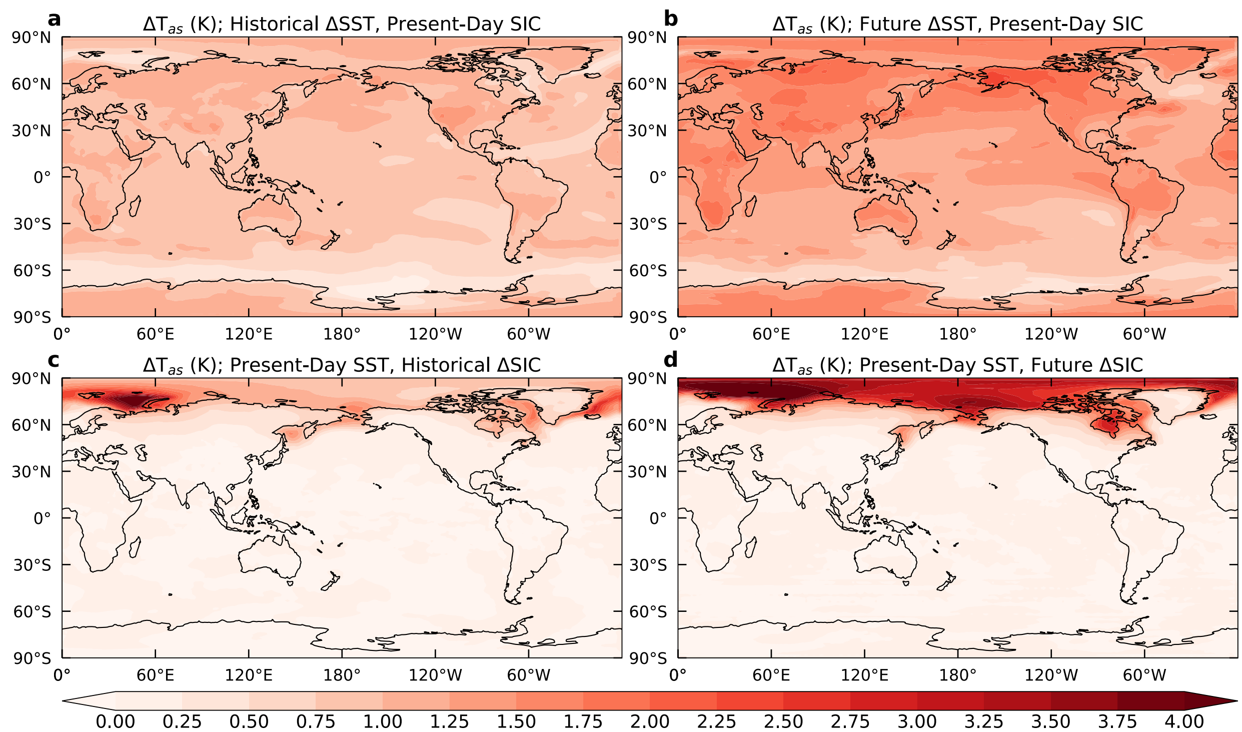

Fig. 2. Multi-model ensemble mean changes in annual-mean surface air temperature (ΔT_{as}) in response to (**a, c**) historical and (**b, d**) future (**a, b**) SST and (**c, d**) SIC changes shown in Fig. 1.

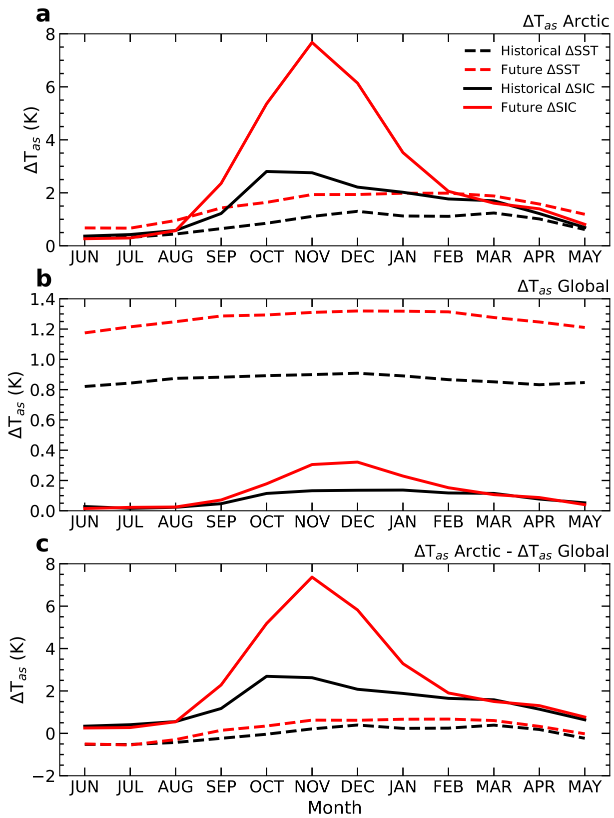

Fig. 3. Multi-model ensemble mean seasonal cycle of surface air temperature changes (ΔT_{as}; in K) in response to historical (black lines) and future (red lines) SST (dashed lines) and SIC (solid lines) perturbations shown in Fig. 1 averaged over the (**a**) Arctic (67°-90°N) and (**b**) globe, and (**c**) Arctic minus global-mean difference (i.e., Arctic amplification).

3.2 Surface energy budget response to Global SST or local Arctic SIC changes

Increased upward surface energy fluxes over sea-ice retreat areas have been shown to drive large Arctic warming and AA in winter (Deser et al. 2010; Boeke and Taylor 2018; Taylor et al. 2018; Dai et al. 2019). In response to SST warming with fixed SIC, we find little change in the net surface energy flux, net surface SW, SH, and LH fluxes over the Arctic Ocean throughout the year (Fig. 4). The upward net surface LW flux decreases by ~1 W m^{-2} for both the historical and future SST warming cases with fixed SIC (Fig. 4c). This represents a small increase in the downward LW radiation, likely due to increased water vapor and enhanced atmospheric energy convergence into the Arctic, rather than changes to surface conditions, as shown below. The suppressed Arctic surface warming and weak oceanic energy flux response to SST warming without SIC changes is consistent with Dai et al. (2019), who found similar results in model simulations with increasing CO_2 concentrations and fixed Arctic sea-ice in flux calculations.

Arctic sea-ice loss greatly influences the magnitude and seasonal cycle of the Arctic oceanic heat flux. From May-August, oceanic *absorption* of energy increases by ~6-12 W m^{-2} in response to historical and future SIC loss (Fig. 4f) while during October-March oceanic *release* of energy increases by ~12-18 W m^{-2} (Fig. 4a). Most of the increased oceanic energy absorption from May-August is due to increased absorption of SW radiation (Fig. 4b), with negligible changes in net surface LW, SH, and LH fluxes (Fig. 4c-e) during summer. In contrast, net surface LW, SH, and LH fluxes are the main contributors to the enhanced cold-season oceanic energy release in response to Arctic sea-ice loss. Specifically, the SH (LH) flux contributes ~4-6 W m^{-2} (~3-4 W m^{-2}) and ~8-10 W m^{-2} (~8-9 W m^{-2}) in response to historical and future SIC loss, respectively (Fig. 4d, e). Further, the ocean surface emits ~1-2 W m^{-2} (~1-3 W m^{-2}) more LW radiation to the atmosphere in autumn and winter in the historical (future) Arctic sea-ice loss runs (Fig. 4c). The large increases in upward surface energy fluxes in response to sea-ice loss play an important role in enhancing warming of the surface air and AA during winter (Fig. 3a).

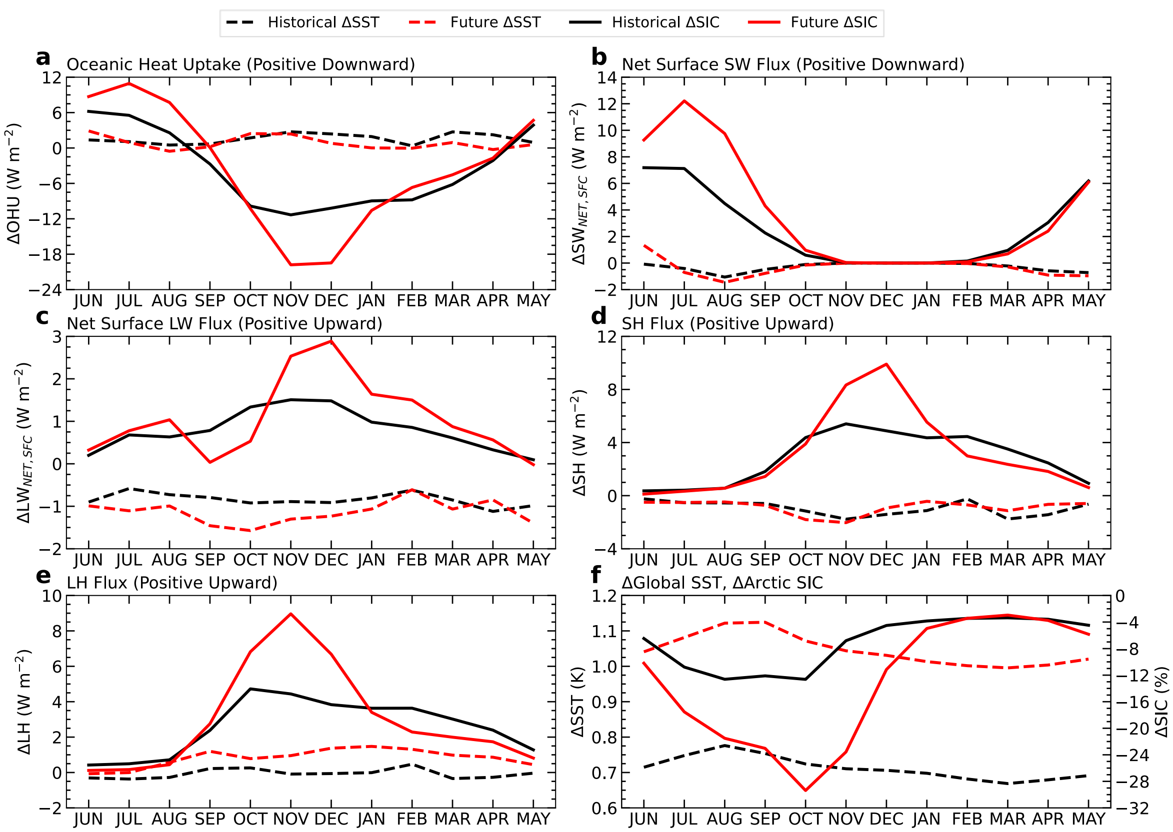

Fig. 4. Arctic (67°-90°N) multi-model ensemble mean seasonal cycle of changes in (**a**) OHU (positive downward), (**b**) net surface SW flux (positive downward), (**c**) net surface LW flux

(positive upward), (**d**) SH flux (positive upward), and (**e**) LH flux (positive upward) in response to historical (black lines) and future (red lines) SST (dashed lines) and SIC (solid lines) perturbations shown in Fig. 1. All values are in W m^{-2} and land surfaces are excluded from averages. (**f**) The seasonal cycle of the historical (black lines) and future (red lines) global SST changes (left y-axis; dashed lines) and Arctic SIC loss (right y-axis; solid lines) specified in the SIC and SST perturbation experiments.

3.3 Feedback seasonal cycles and warming contributions

The contrasting surface warming responses to global SST changes or local sea-ice loss greatly influence Arctic climate feedbacks and atmospheric energy convergence changes. Under global SST warming with fixed SIC, Arctic atmospheric energy convergence (Fig. 5f) and water vapor feedback (Fig. 5b) become important contributors to the Arctic TOA flux change. Specifically, atmospheric energy convergence into the Arctic responds similarly to historical and future SST warming, with increases of ~4-7 W m^{-2} during the winter months and ~2-4 W m^{-2} in summer (Fig. 5f). This suggests that without changes in sea ice, increased atmospheric energy transport becomes an important contributor to small cold season Arctic warming and AA (Fig. 3). Further, the magnitude and seasonal cycle of the water vapor feedback is similar between the historical and future SST cases, with maximum water vapor feedback (1.5-1.8 W m^{-2}) from May-August and minimum water vapor feedback (0.4-0.7 W m^{-2}) during October-March (Fig. 5b). This is expected as the warm-season Arctic would see larger water vapor increases due to its warmer mean air temperatures. Arctic surface albedo (Fig. 5a), lapse rate (Fig. 5c), and Planck (Fig. 5d) feedbacks weakly respond to SST increases without sea-ice loss. Lastly, we note that the net cloud feedback produces slight cooling (-1.5~-2.0 W m^{-2}) in response to SST increases for June-August (Fig. 5e).

In response to sea-ice loss, Arctic surface albedo feedback increases by ~7-8 W m^{-2} and ~12-18 W m^{-2} for historical and future cases during the sunlit months (i.e., April-September) due to increased exposure of dark water surfaces (Fig. 5a). The ocean, rather than the atmosphere, absorbs much of the extra SW radiation (Fig. 4a), resulting in weak summer surface warming (Fig. 3a). Cloud feedback is negative in response to sea-ice loss during April-August, and the cooling is larger in the future SIC case (-1.5~-4.5 W m^{-2}) than the preindustrial SIC run (-1.0~-1.5 W m^{-2}). Lapse rate (Fig. 5c) and Planck (Fig. 5d) feedbacks weakly respond to historical or future Arctic SIC changes in summer due to small surface warming (Fig. 3a) during the sunlit season. We also find negligible water vapor feedback in response to Arctic sea-ice loss throughout the year, which differs from the noticeable water vapor feedback in response to SST warming (Fig. 5b).

The large cold-season surface warming in response to historical and future Arctic sea-ice loss enhances Arctic lapse rate (Fig. 5c) and Planck (Fig. 5d) feedbacks. When Arctic surface warming (Fig. 3a) and AA (Fig. 3c) peak from October-December, the lapse rate feedback increases the incoming TOA radiative flux by ~4-6 W m^{-2} (~8-11 W m^{-2}) and the Planck feedback opposes warming by -6~-8 W m^{-2} (-16~-20 W m^{-2}) due to historical (future) sea-ice loss. Note that the month of maximum (minimum) lapse rate (Planck) feedback in the historical and future SIC cases (Fig. 5c) corresponds to the month of peak Arctic surface warming (Fig. 3a), which in turn is related to peak oceanic heating (Fig. 4a) induced by sea-ice loss (Fig. 4f) in these simulations. The cloud feedback in response to future Arctic sea-ice loss also enhances the net incoming TOA radiative flux from October-January by ~2.5-3.0 W m^{-2}, but the cloud feedback is weak (<1.0 W m^{-2}) during winter in response to historical sea-ice loss (Fig. 5e). In contrast to the SST change simulations, Arctic atmospheric energy convergence weakens by 7 W m^{-2} and 9-13 W m^{-2} in response to historical and future sea-ice loss from October-January, respectively (Fig. 5f). Enhanced Arctic warming in response to sea-ice loss in the non-summer months (Fig. 3a) weakens the temperature gradient between the midlatitudes and polar regions, thus reducing atmospheric energy convergence into the Arctic region.

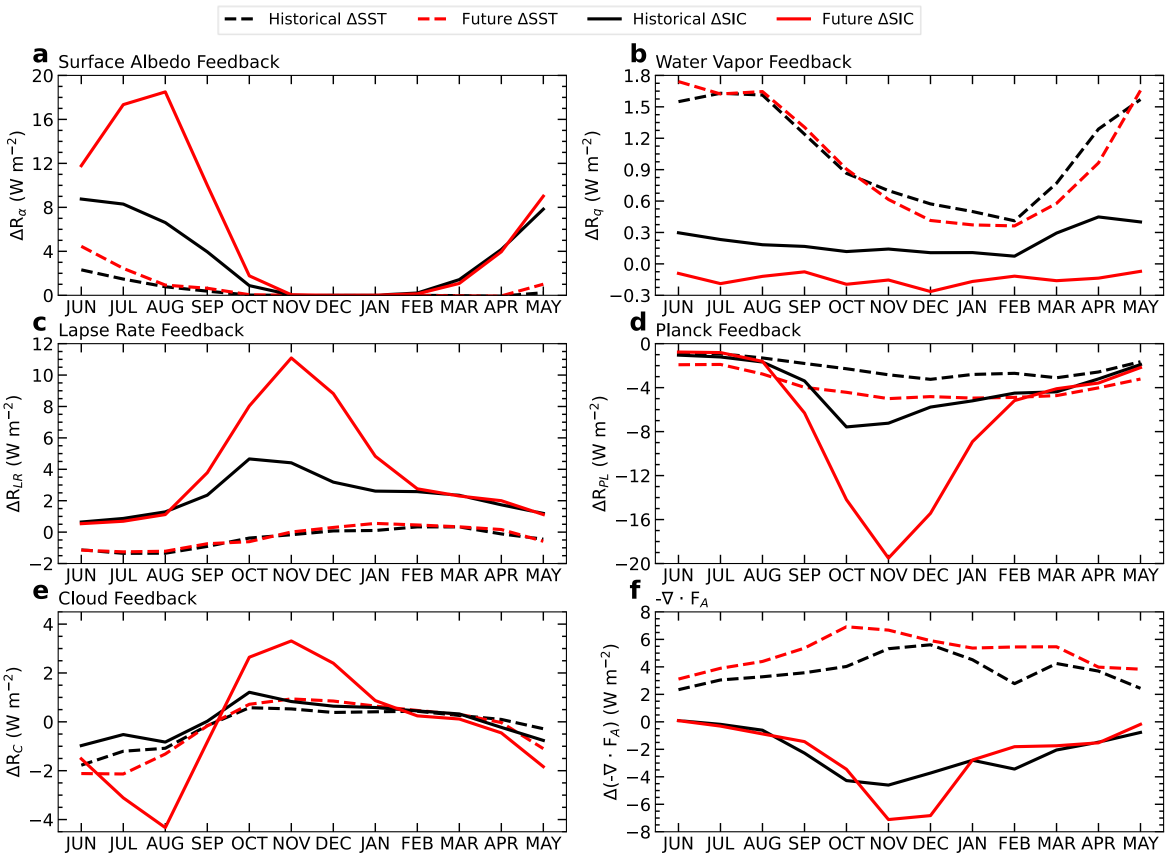

Fig. 5. Arctic (67°-90°N) multi-model ensemble mean seasonal cycle of the (**a**) surface albedo, (**b**) water vapor, (**c**) lapse rate, (**d**) Planck, and (**e**) cloud feedbacks, and (**f**) changes in atmospheric energy convergence into the Arctic in response to historical (black lines) and future (red lines) SST (dashed lines) and SIC (solid lines) changes shown in Fig. 1. All values are in W m^{-2} and land surfaces are excluded in averages except for the case shown in (**f**).

Warmer SSTs enhance poleward atmospheric energy transport at all latitudes for each model for the historical (Fig. 6a) and future (Fig. 6b) SST warming cases, with slightly larger increases in the northern hemisphere than southern hemisphere from October-March. All models, except CESM2, show enhanced cold season northward energy transport with peak increases of ~0.18 (~0.22) PW around ~45°-50°N for the historical (future) SST warming cases. In CESM2, atmospheric energy transport shows peak increases of 0.27 (0.40) PW around 30°N for October-March. Thus, without large Arctic warming related to sea-ice loss, the atmosphere displaces energy surpluses poleward. For the SIC perturbation experiments, there is a net decrease in poleward atmospheric energy transport around 30°-90°N with a maximum decrease of -0.05~-0.12 PW around 60°N but little change south of 30°N for both historical (Fig. 6c) and future (Fig. 6d) sea-ice loss, consistent with Deser et al. (2015). Therefore, SST-induced background warming enhances atmospheric poleward energy transport into the polar regions, while large Arctic warming in response to sea-ice loss weakens atmospheric poleward energy transport over the northern mid-high latitudes.

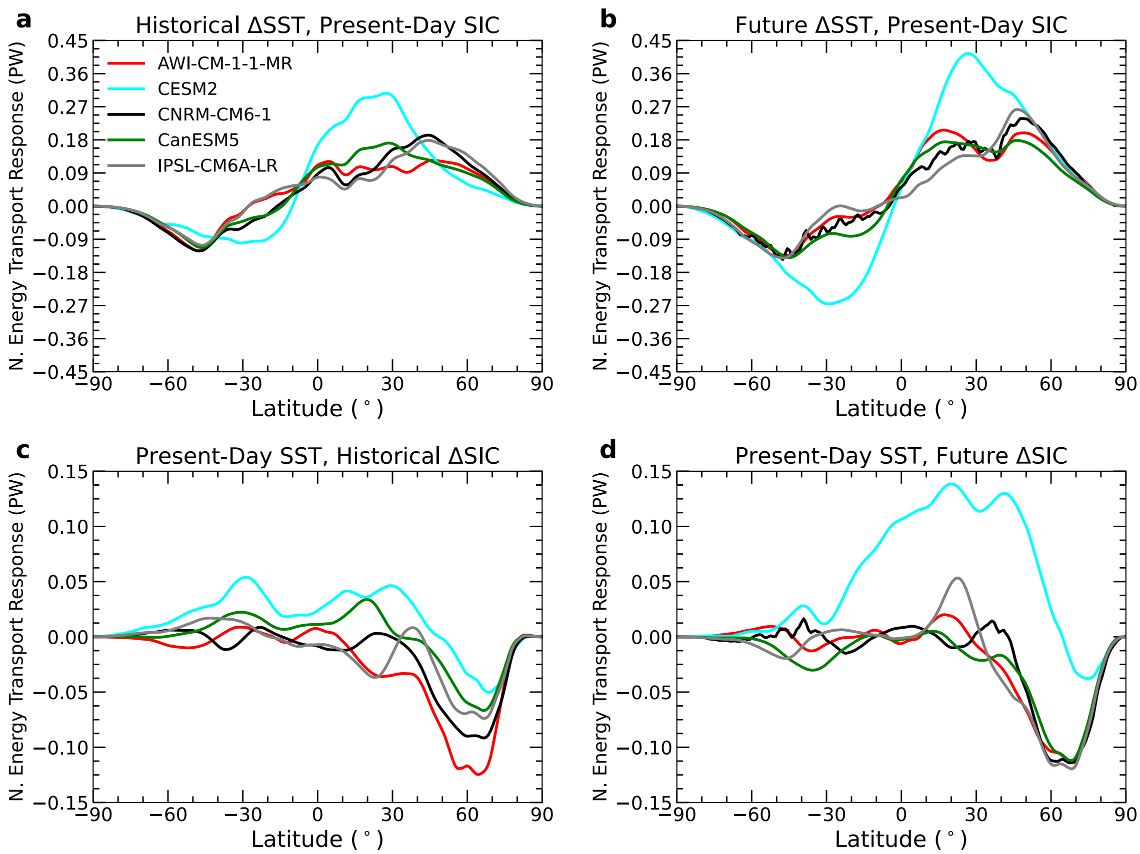

Fig. 6. Changes in October-March mean Arctic northward energy transport in response to (**a**, **c**) historical and (**b**, **d**) future (**a**, **b**) SST and (**c**, **d**) SIC changes shown in Fig. 1.

Figure 7 shows the *potential* warming contributions of the climate feedbacks over the Arctic and the tropics. A larger (smaller) Arctic than tropical warming contribution indicates that the process contributes positively (negatively) to AA (Pithan and Mauritsen 2014). We focus on the warming contributions from October-March (rather than the annual-mean) as surface warming and AA is largest during the cold season (Fig. 3a) and warm-season positive contributions (such as that from surface albedo feedback) do not lead to significant surface warming as the energy is stored in the upper ocean. Atmospheric energy convergence is the largest contributor to cold-season AA under historical (Fig. 7a) and future (Fig. 7b) global SST warming, as it redistributes the energy from the lower latitude oceans, where SSTs increase, to the Arctic region. In contrast, oceanic heat release opposes AA in response to global SST warming (Fig. 7a-b) because the warmer SSTs produce a greater ocean-to-atmosphere energy flux outside the Arctic, thus causing more warming in the tropics than in the Arctic. Water vapor feedback makes a small contribution to Arctic warming due to low October-March mean temperatures but contributes to ~1 K of warming in the tropics in response to global SST warming (Fig. 7b), opposing AA. Without sea-ice loss, lapse rate feedback contributes little to Arctic warming but produces weak tropical cooling in response to historical (Fig. 7a) and future (Fig. 7b) SST increases from October-March. The local Planck feedback (relative to the global-mean Planck feedback) slightly contributes to AA in the SST warming runs because the cooling effects from Planck feedback are slightly less in the Arctic region than over the rest of the world (Fig. 7a-b). Surface albedo and cloud feedbacks contribute little to Arctic warming or AA in response to global SST increases and fixed Arctic SIC during October-March for historical (Fig. 7a) and future (Fig. 7b) cases.

In response to Arctic sea-ice loss with fixed global SSTs, oceanic heat release is the largest contributor to AA from October-March in historical (Fig. 7c) and future (Fig. 7d) SIC cases, followed by the positive lapse rate feedback. This supports previous studies that showed that sea-ice loss and oceanic energy release during Arctic winter are necessary to trigger large surface warming and thus strong positive lapse rate feedback in the Arctic (Feldl et al. 2020; Jenkins and Dai 2021; Dai and Jenkins 2023). The local Planck feedback (relative to the global-mean Planck feedback) also contributes to Arctic warming and AA in response to historical (Fig. 7c) and future (Fig. 7d) Arctic SIC changes by cooling the Arctic region less than the tropics. Additionally, positive cloud feedback makes a slight contribution to cold-season Arctic warming and AA in response to future Arctic SIC loss (Fig. 7d), but the contribution is negligible in the historical SIC loss run (Fig. 7c). Water vapor feedback is suppressed over the Arctic and globe in the historical (Fig. 7c) and future (Fig. 7d) SIC runs, suggesting that local sea-ice loss and water vapor feedback are decoupled, as found previously (Jenkins and Dai 2021). In contrast to the perturbed SST runs, the atmosphere displaces energy away from the Arctic in response to cold season sea-ice loss (Fig. 7c-d), thus opposing AA.

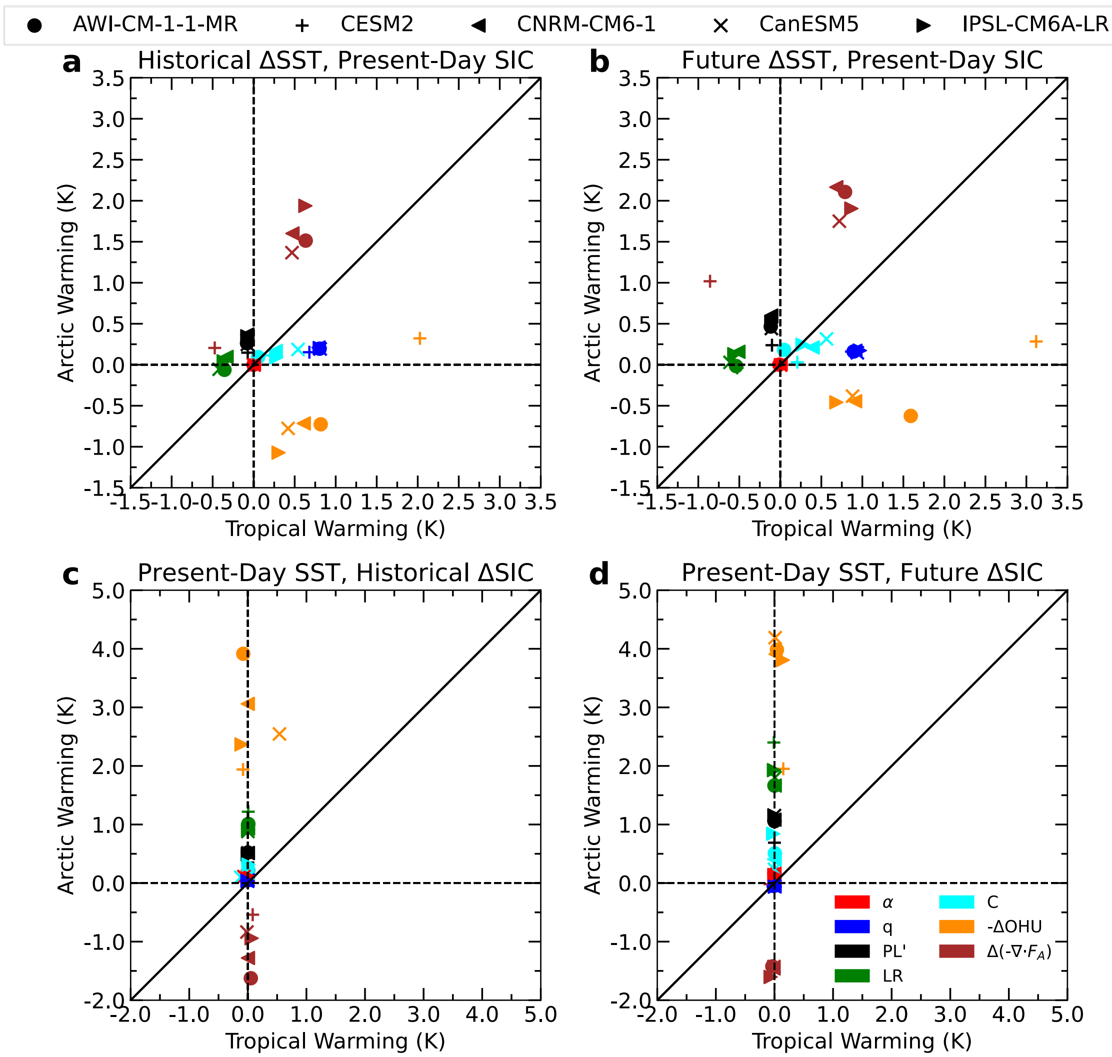

Fig. 7. Inter-model spread in ensemble mean, October-March *potential* warming contributions (in K) of Arctic (67°-90°N) vs. tropical (23.5°S-23.5°N) surface albedo (α), water vapor (q), Planck (PL'), lapse rate (LR), and cloud (C) feedbacks, and changes in oceanic heat release (-ΔOHU; positive upwards), and atmospheric energy convergence ($\Delta(-\nabla \cdot \boldsymbol{F_A})$) in response to (**a**, **c**) historical and (**b**, **d**) future (**a**, **b**) SST and (**c**, **d**) SIC perturbations shown in Fig. 1.

3.4 Physical processes underlying climate feedbacks

Water vapor feedback is complicated in high latitudes due to local temperature inversions and low amounts of water vapor (Curry et al. 1995; Sejas et al. 2018). Global maps reveal that SST warming (Fig. 8a, b) has a larger effect than local sea-ice loss (Fig. 8c, d) on water vapor feedback in both the Arctic and remote areas. Specifically, water vapor feedback is largest near the equator at ~2-5 W m^{-2} in response to historical (Fig. 8a) and future (Fig. 8b) SST warming and decreases poleward to ~0.5-1.0 W m^{-2} in the Arctic region (Fig. 8a, b). The cold-season water vapor feedback is weak in response to Arctic sea-ice loss (Fig. 8c, d), including over the Arctic where low-level specific humidity increases (Fig. 9c, d). This is due to low or negative values of the October-March LW and net (i.e., LW+SW) water vapor kernel in the Arctic lower troposphere (Fig. 10a, c). Because the water vapor feedback is most sensitive to upper tropospheric water vapor content (Shell et al. 2008; Soden et al. 2008; Pendergrass et al. 2018), the low-level water vapor increases in response to Arctic sea-ice loss do not lead to large TOA flux changes.

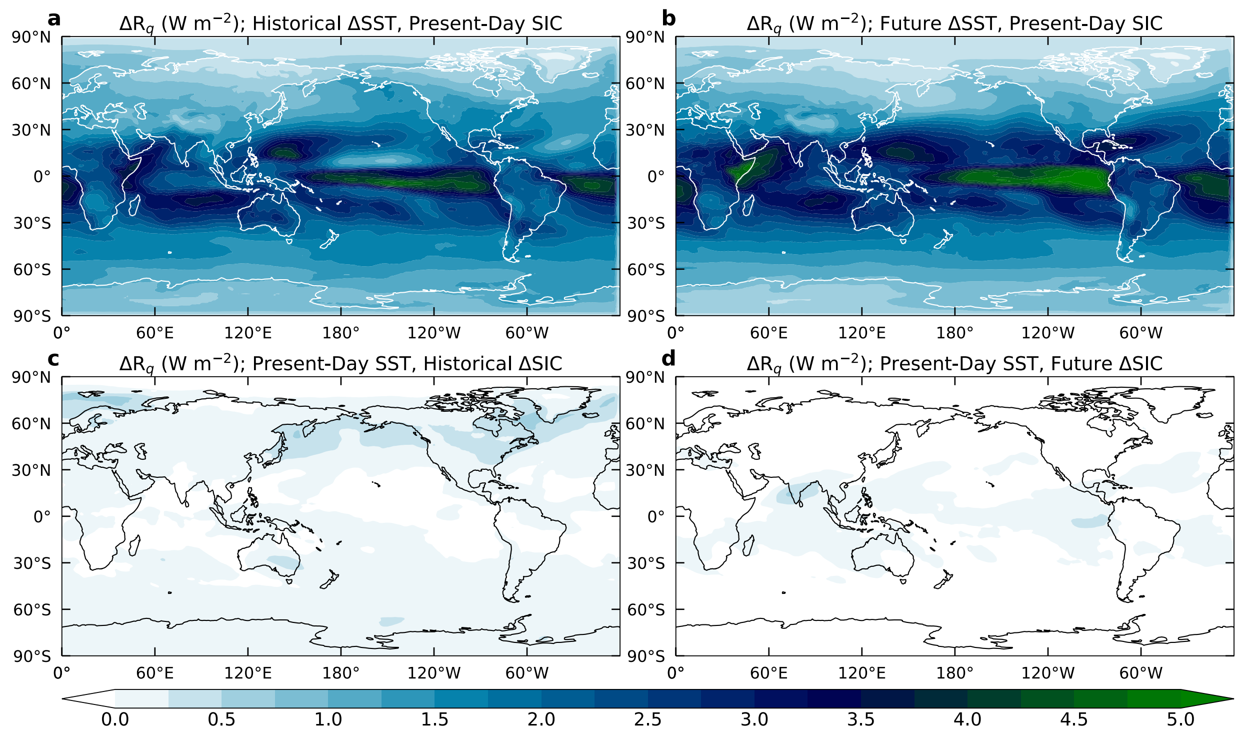

Fig. 8. Multi-model ensemble mean October-March water vapor feedback (in W m^{-2}) in response to (**a, c**) historical and (**b, d**) future (**a, b**) SST and (**c, d**) SIC changes shown in Fig. 1.

Slight positive water vapor feedback occurs over sea-ice loss areas in the historical SIC loss run (~0.50-0.75 W m^{-2}; Fig. 8c) but there are negligible water vapor feedback effects in the Arctic under future SIC conditions (Fig. 8d). As the October-March LW and net water vapor kernel is negative near the surface (Fig. 10a, c), any increase in moisture in the lower troposphere will result in enhanced radiative emission to space (i.e., a negative water vapor radiative effect). In response to future Arctic SIC (Fig. 9d), there are greater increases in the natural logarithm of specific humidity [Δln(q)] in the lower troposphere than in the historical case (Fig. 9c). Thus, greater future lower tropospheric moistening in the Arctic region produces a more negative water vapor radiative effect at the TOA. We also note that there is a large spread (as shown by the standard deviation) among the PAMIP models and individual ensemble members in upper tropospheric moistening in the perturbed Arctic SIC runs, where there is little change in the mean Δln(q) (Fig. 10c, d). Thus, some ensemble members may have experienced a slight decrease in upper tropospheric Δln(q) in response to Arctic sea-ice loss with fixed global SST, enhancing outgoing LW radiation at the TOA. In contrast, the historical (Fig. 10a) and future (Fig. 10b) perturbed SST runs experienced slightly greater Δln(q) in the upper troposphere than the lower troposphere for both warm and cold seasons. Due to positive values of the TOA LW and net Arctic water vapor kernel in the upper troposphere (Fig. 10a, c), top-heavy moistening in response to global SST warming produces a positive water vapor feedback from the TOA perspective.

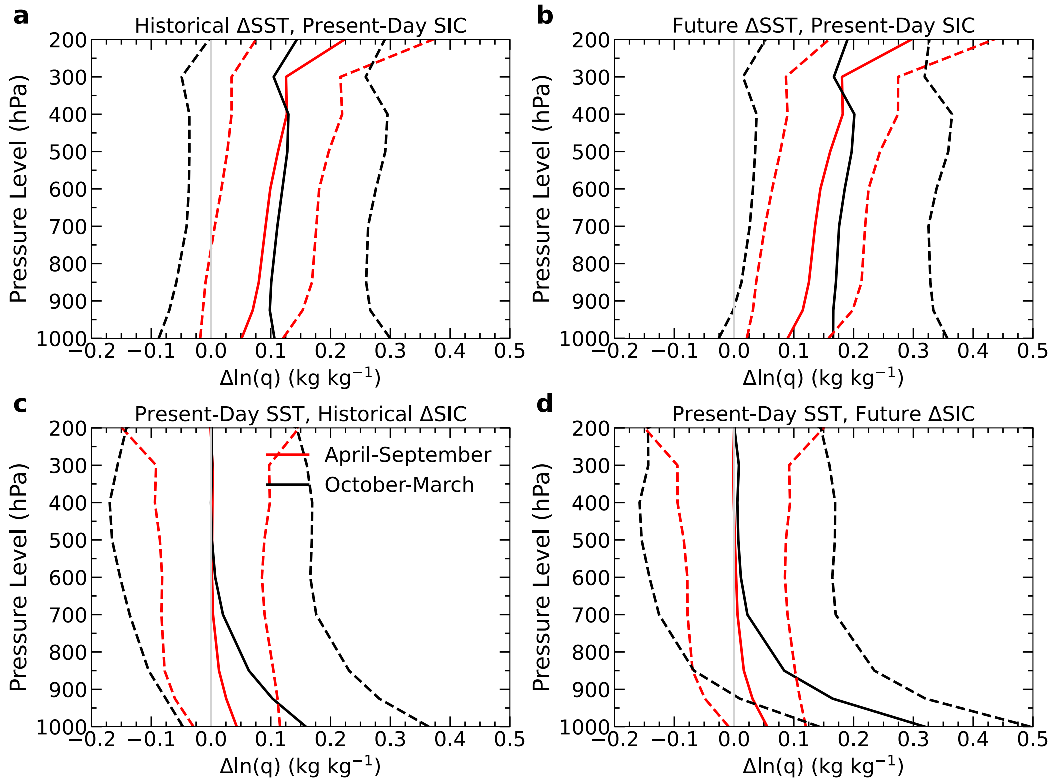

Fig. 9. Multi-model, ensemble mean (solid lines) Arctic (67°-90°N; land surfaces excluded) changes in the natural logarithm of specific humidity (in kg kg^{-1}; solid lines) in response to (**a, c**) historical and (**b, d**) future (**a, b**) SST and (**c, d**) SIC changes shown in Fig. 1. Dashed lines show ±1 standard deviation from the multi-model ensemble mean profile.

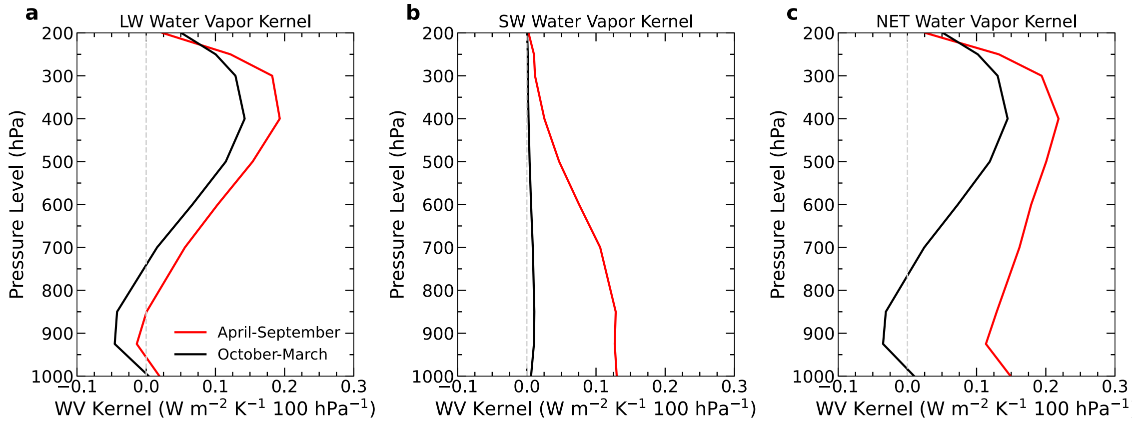

Fig. 10. Profiles of the Pendergrass et al. (2018) TOA (**a**) LW, (**b**) SW, and (**c**) NET (LW+SW) water vapor kernel (in W m^{-2} K^{-1} 100 hPa^{-1}) averaged over the Arctic region (67°-90°N).

Arctic low cloud amount has been suggested to increase during the cold season in response to sea-ice loss due to decreased lower tropospheric stability (Kay and Gettelman 2009; Jenkins et al. 2023), thus affecting Arctic cloud feedback (Vavrus 2004; Morrison et al. 2019; Jenkins and Dai 2022). We find weak October-March cloud feedback in response to perturbed SST with fixed Arctic SIC for historical (Fig. 11a) and future (Fig. 11b) cases, suggesting that remote processes do not greatly impact Arctic cloud feedback. On the other hand, Arctic sea-ice loss produces a large positive cloud feedback response in winter, especially in regions with large sea-ice loss and surface warming (Fig. 11c, d). For the run with historical SIC loss, cloud feedback enhances the TOA radiative flux by ~2-5 W m^{-2} in the Barents-Kara Seas region and by ~0.5-1.0 W m^{-2} in the Chukchi Sea, where the largest sea-ice loss and surface warming occurs. Under future Arctic sea-ice loss, cold-season cloud feedback is largest in the Barents-Kara Seas (~3-5 W m^{-2}) except the warming effects from clouds extend into the Central Arctic Ocean. This is likely related to the greater area with large sea-ice loss (Fig. 1b, d) and surface warming (Fig. 2c-d) in the future case than in the historical case.

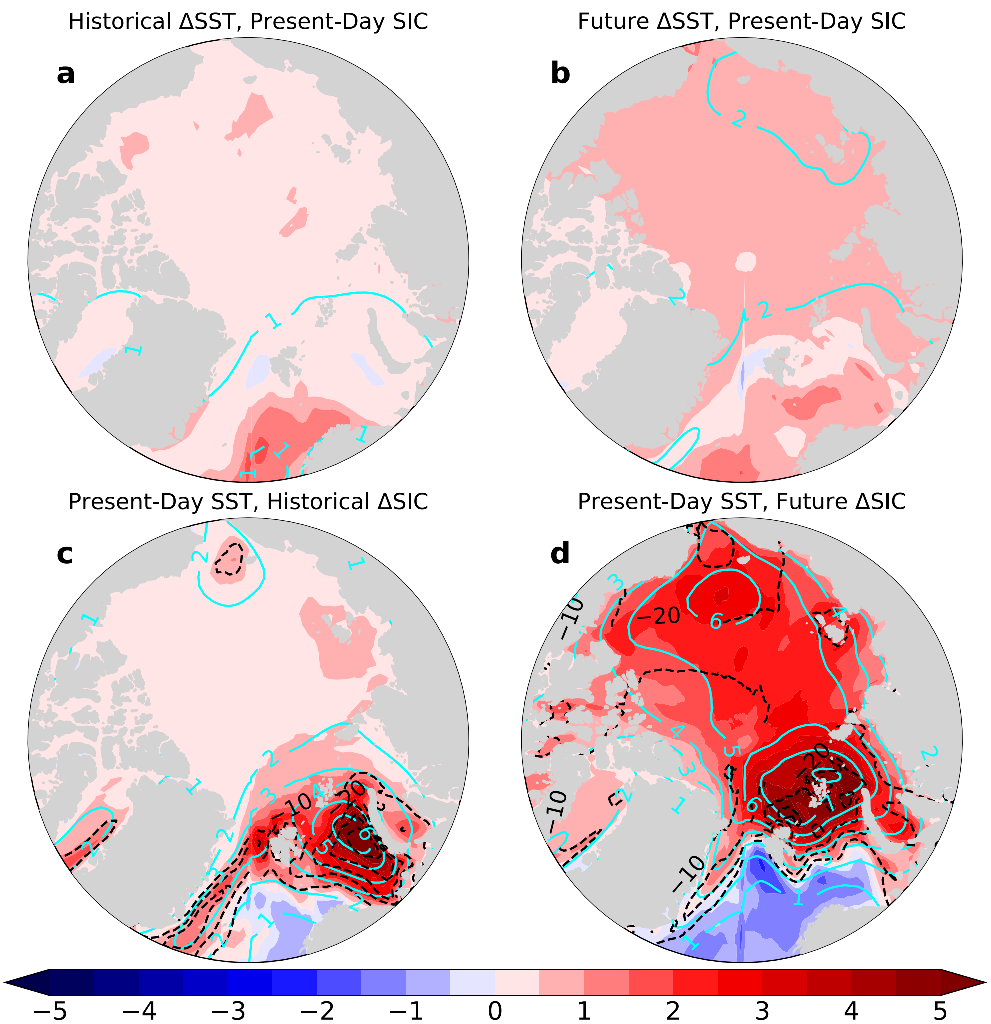

Fig. 11. Multi-model ensemble mean TOA radiative flux change due to the cloud feedback (shading; in W m^{-2}) and change in surface air temperature (cyan contours; in K) averaged over October-March in response to (**a, c**) historical and (**b, d**) future (**a, b**) SST and (**c, d**) SIC changes. Black contours in (**c**) and (**d**) show the change in Arctic SIC for October-March.

The lapse rate feedback experiences large seasonal and spatial variations in the Arctic in response to SST warming or Arctic SIC loss. From October-March, the lapse rate feedback is negative-neutral in response to the global SST warming (Fig. 12a, b) due to relatively uniform vertical warming profiles (Fig. 13a, b). We note that without changes in SIC, there are negligible changes in Arctic oceanic heat uptake or surface warming in the cold season, leading to suppressed lapse rate feedback (Fig. 12a, b). In contrast, cold-season sea-ice loss enhances Arctic lapse rate feedback for historical (Fig. 12c) and future (Fig. 12d) SIC cases when surface and lower tropospheric warming outpaces warming in the mid-upper troposphere (Fig. 13c, d). We note that lapse rate feedback strengthens (~6-10 W m^{-2}) in regions with the greatest October-March oceanic heat release and surface warming in response to historical (Fig. 12c) and future (Fig. 12d) sea-ice loss, consistent with previous studies (Dai et al. 2019; Feldl et al. 2020; Boeke et al. 2021; Jenkins and Dai 2021, 2022; Dai and Jenkins 2023). Thus, sea-ice loss is necessary to produce bottom-heavy warming and trigger Arctic positive lapse rate feedback during winter, as shown previously by Dai and Jenkins (2023) using coupled model experiments.

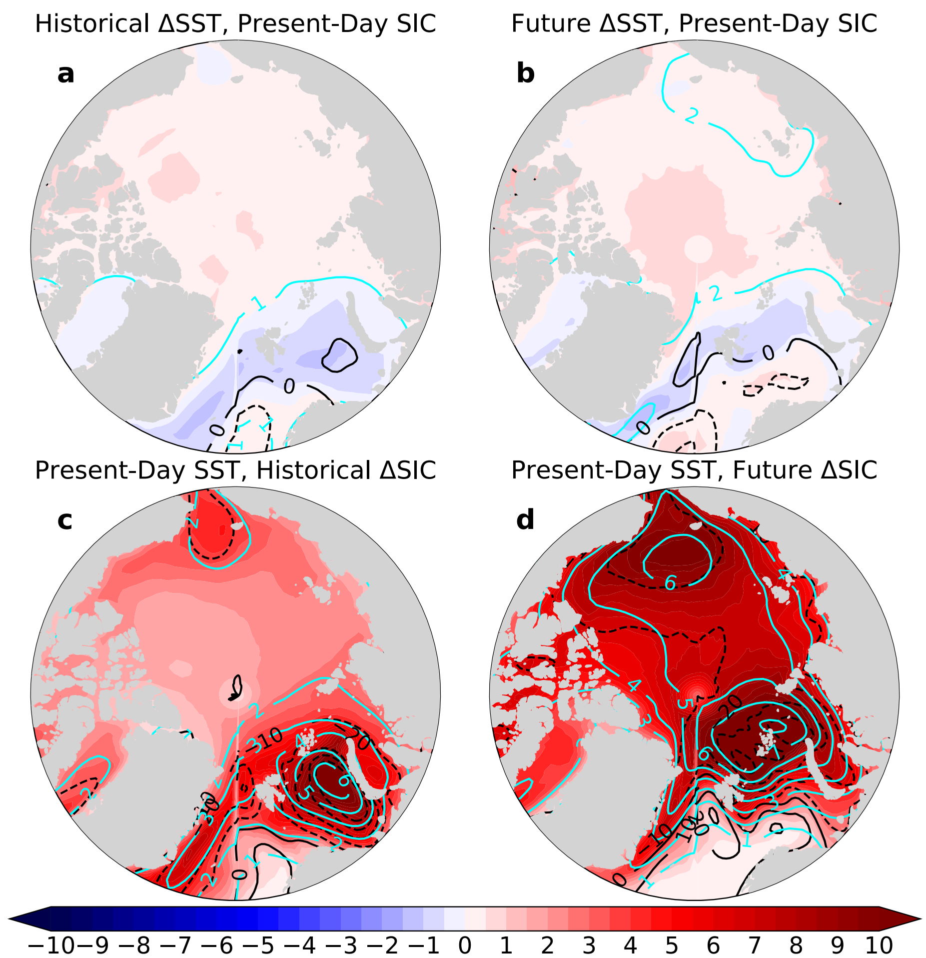

Fig. 12. Multi-model, ensemble mean TOA radiative flux change due to the lapse rate feedback (shading; in W m^{-2}), changes in oceanic heat uptake (black contours; in W m^{-2}; positive downward), and changes in surface air temperature (cyan contours; in K) averaged over October-March in response to (**a**, **c**) historical and (**b**, **d**) future (**a**, **b**) SST and (**c**, **d**) SIC changes.

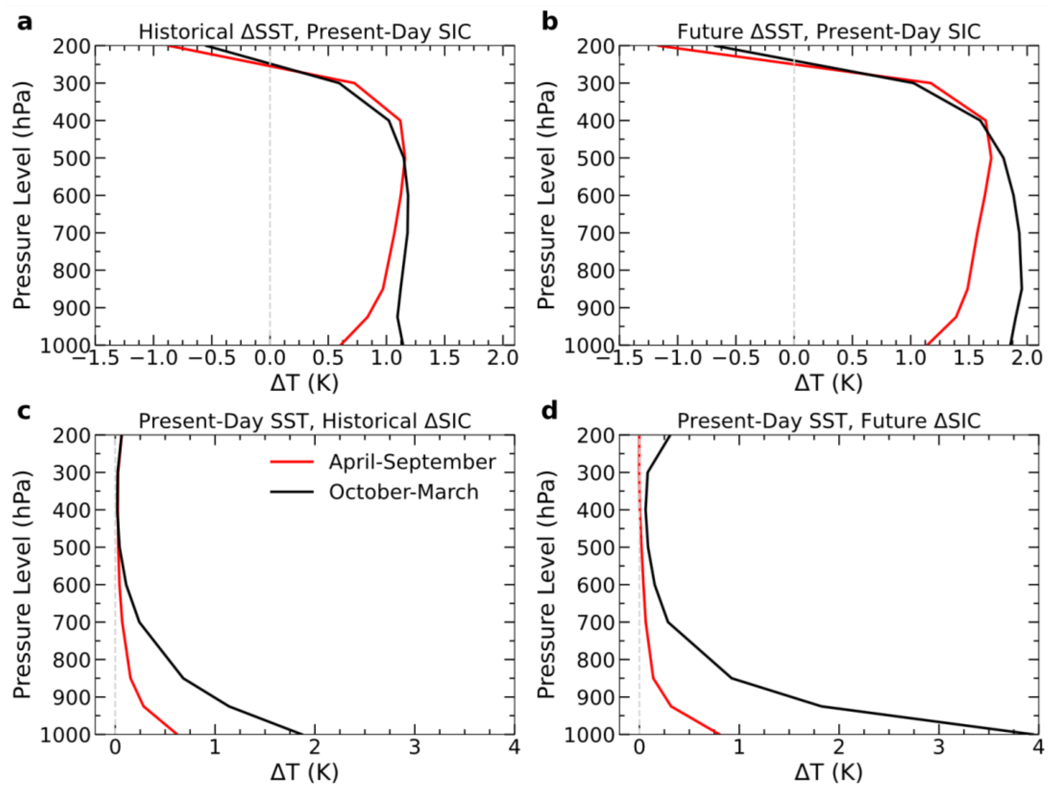

Fig. 13. Multi-model, ensemble mean Arctic (67°-90°N; land surfaces excluded) temperature change profiles (in K) averaged over April-September (red lines) and October-March (black lines) in response to the (**a**) historical and (**b**) future global SST warming, and (**c**) historical and (**d**) future Arctic sea-ice loss.

3.5 Response to simultaneous SST and SIC changes

We compare the Arctic vs. tropical October-March potential warming contributions of climate feedbacks, changes in atmospheric energy convergence and oceanic heat release in response to historical global SST warming and historical polar sea-ice loss together (i.e., pdSST-pdSIC minus piSST-piSIC; Fig. 14a; referred to as TOTAL) and the sum of the separate responses to historical SST warming (i.e., pdSST-pdSIC minus piSST-pdSIC) and historical polar sea-ice loss (i.e., pdSST-pdSIC minus pdSST-piArcSIC and pdSST-piAntSIC) (Fig. 14b; referred to as SUM). The warming contributions of the lapse rate, water vapor, cloud, and Planck feedbacks in TOTAL match SUM well, with the lapse rate feedbacks making the largest contribution to AA (Fig. 14). Except for CESM2 in TOTAL, the change in atmospheric energy convergence makes roughly equal warming contributions to Arctic and tropical warming from October-March, suggesting that remote SST warming and Arctic sea-ice loss have opposing effects on the horizontal atmospheric energy flux. The oceanic heat release changes in IPSL-CM6A-LR makes a greater contribution to Arctic than tropical warming, but there are slight discrepancies between CanESM5 and CESM2 oceanic heat release between TOTAL and SUM. In TOTAL, CanESM5 and CESM2 oceanic heat release changes contributes roughly the same amount to Arctic and tropical warming; however, CESM2 (CanESM5) produces slightly greater Arctic (tropical) warming in SUM. The surface albedo feedback is inactive from October-March due to lack of sunlight and is not a major contributor to large cold-season AA.

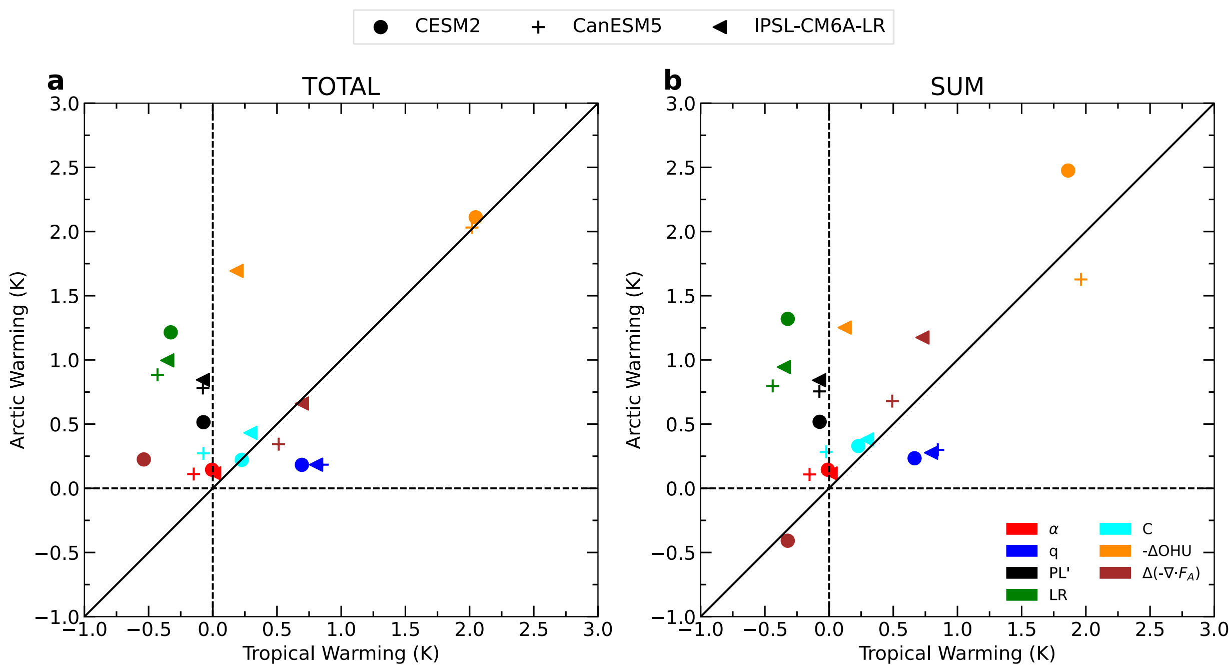

Fig. 14. Inter-model spread in the ensemble mean October-March *potential* warming contributions (in K) for Arctic (67°-90°N) and tropical (23.5°S-23.5°N) surface albedo (α), water vapor (q), Planck (PL'), lapse rate (LR), and cloud (C) feedbacks, and changes in oceanic heat release (-ΔOHU; positive upwards) and atmospheric energy convergence ($\Delta(-\nabla \cdot \boldsymbol{F_A})$) in response to historical changes in global SST and polar SIC for (**a**) TOTAL (i.e., global SST and polar SIC change together) and (**b**) SUM (i.e., sum of the response to the SST and SIC change separately).

The northward atmospheric energy transport response to the SST and SIC perturbations is similar among TOTAL (Fig. 15a) and SUM (Fig. 15b), with little difference between the two cases (Fig. 15c). In the tropical regions (i.e., 30°S-30°N), remote SST warming enhances poleward atmospheric energy transport by ~0.1-0.15 PW in the southern hemisphere and ~0.1-0.35 PW in the northern hemisphere. Around 60°-90°N, there is little net change in atmospheric energy transport in response to simultaneous SST and SIC changes, suggesting that remote warming due to SST changes and local Arctic warming related to sea-ice loss have opposing effects on Arctic atmospheric energy transport. The similarity of climate feedbacks (Fig. 14) and the atmospheric energy transport (Fig. 15) response between TOTAL and SUM suggest that the effects of SIC or SST changes can be linearly separated. In other words, the individual responses to SST or SIC perturbations approximately sum to the combined influence of changes in SST and SIC.

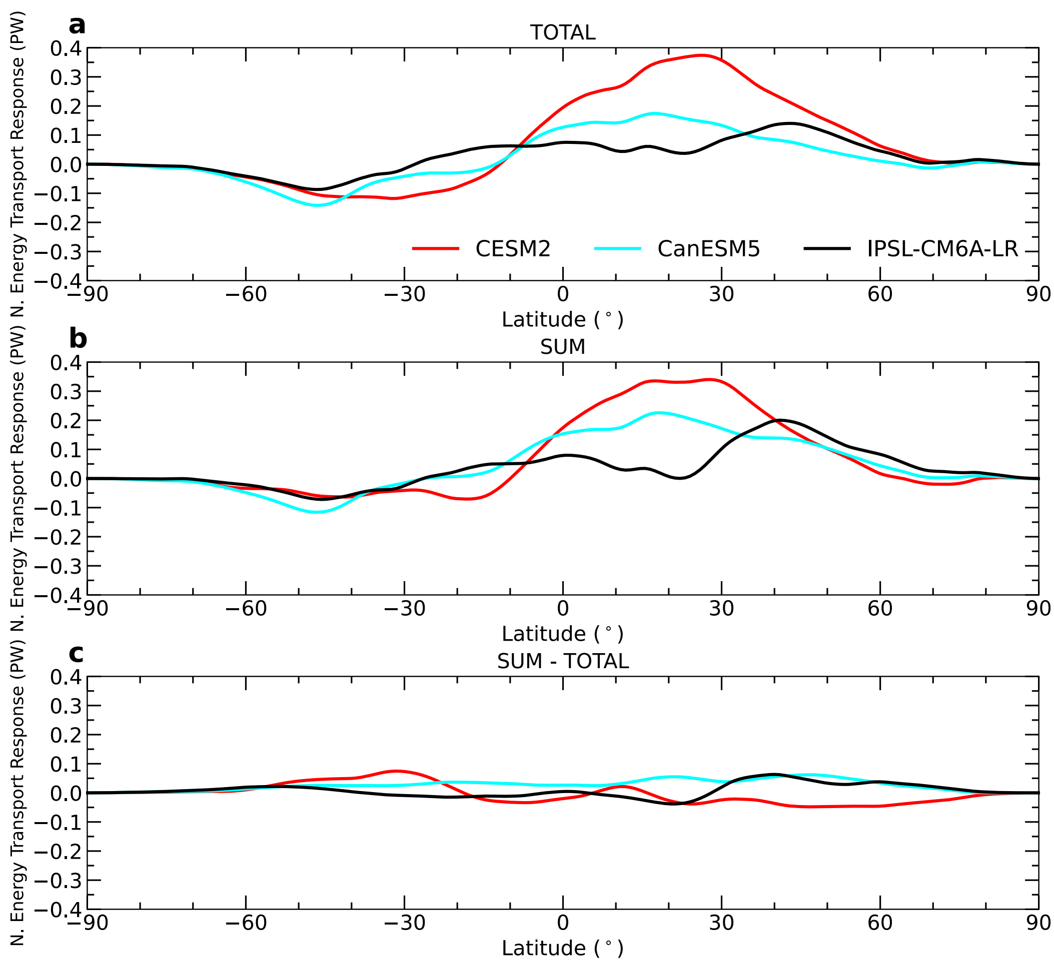

Fig. 15. October-March Arctic northward energy transport response (in PW) in response to historical changes in SST and SIC for (**a**) TOTAL and (**b**) SUM, and (**c**) difference between (**b**) and (**a**).

4. Summary and Conclusions

We investigated the impacts of historical and future Arctic sea-ice loss and global SST increases on Arctic climate feedbacks, atmospheric energy convergence into the Arctic, and oceanic heat release using PAMIP atmosphere-only simulations. The SST increase with fixed polar sea ice results in relatively uniform global warming with negligible AA for both historical and future cases. In contrast, historical and future Arctic sea-ice loss leads to large Arctic warming with negligible effects south of ~60°N, although this may not be the case in fully coupled simulations (Deser et al. 2015). The PAMIP experiments allowed us to separate the response of Arctic climate feedbacks, atmospheric energy convergence, and oceanic heat release to background global warming without AA (as in the SST perturbation runs) or to large AA with negligible warming outside the Arctic (as in the SIC change runs). We also found striking similarities between the historical simulations with both SST and SIC changes together (i.e., TOTAL), and the sum of the individual responses to the historical SST and polar SIC changes (i.e., SUM) in terms of Arctic climate feedbacks and atmospheric energy transport response.

Under warmer global SSTs without sea-ice loss, Arctic winter oceanic heat release is suppressed leading to weak Arctic cold season warming. Instead, enhanced poleward atmospheric energy convergence rather than increased oceanic heat release becomes the dominant contributor to *small* AA in response to global SST increases with fixed Arctic sea-ice. We also found strong global water vapor feedback in the historical and future SST warming runs, especially in the tropics. However, the combined effects of enhanced atmospheric energy convergence into the Arctic and positive water vapor feedback produce weak Arctic warming without large sea-ice loss and enhanced oceanic heat release. We also found that under global SST warming with fixed Arctic SIC, the Arctic experiences vertically uniform or top-heavy warming, producing a neutral or negative lapse rate feedback. Thus, the lapse rate feedback does not make a large contribution to Arctic warming or AA without the bottom-heavy warming effects of enhanced oceanic energy release due to sea-ice loss. Lastly, Arctic cloud and surface albedo feedbacks responded weakly to warmer global SST with fixed Arctic SIC in the historical and future cases.

In contrast, retreating sea ice produces strong bottom-heavy warming and moistening in autumn and winter due to enhanced oceanic energy release in regions with newly exposed water surfaces, as shown in previous studies (Deser et al. 2010; Screen and Simmonds 2010a, b; Boeke and Taylor 2018; Dai et al. 2019; Dai and Jenkins 2023). Strong lower tropospheric warming enhances Arctic positive lapse rate feedback, which greatly contributes to AA during the cold season (e.g., Jenkins and Dai 2021; Dai and Jenkins 2023). Additionally, bottom-heavy moistening in response to Arctic sea-ice loss has little impact on the TOA radiative flux due to its low sensitivity to lower tropospheric water vapor (Shell et al. 2008; Soden et al. 2008; Pendergrass et al. 2018). Instead, enhanced moistening in the mid-upper troposphere, as in the SST warming runs, increases the Arctic TOA radiative forcing by increasing water vapor's LW absorption in the upper troposphere. Arctic surface albedo feedback activates during the sunlit season in response to sea-ice loss but does not significantly raise surface temperatures in summer. We also find reduced poleward atmospheric energy transport in the northern hemisphere mid-high latitudes due to historical and future Arctic sea-ice loss with fixed global SST, consistent with Hahn et al. (2023).

Many previous studies have quantified Arctic climate feedback processes in fully coupled model simulations and have suggested that lapse rate feedback is a dominant contributor to AA by comparing its Arctic and tropical warming contributions (Pithan and Mauritsen 2014; Stuecker et al. 2018; Feldl et al. 2020; Hahn et al. 2021). Here, we showed that Arctic lapse rate feedback weakens without sea-ice loss as in the global SST warming with fixed Arctic SIC simulations and activates in the simulations with reduced Arctic sea-ice and enhanced oceanic energy release during the cold season. We emphasize that sea-ice loss, enhanced oceanic heating, and bottom-heavy warming trigger Arctic positive lapse rate feedback, but the effects from the lapse rate feedback in turn amplify Arctic warming resulting from the oceanic energy release, as shown previously (Dai and Jenkins 2023). Therefore, the coupled sea ice loss-lapse rate feedback relationship is an important process underlying large cold-season AA. Further, our results are qualitatively consistent with Jenkins and Dai (2021) and Dai and Jenkins (2023), who showed suppressed lapse rate feedback in a coupled model simulation with fixed sea ice and increasing CO_2.

The surface albedo feedback has been suggested to be an important process underlying AA (Hall 2004; Winton 2006) due to its large annual-mean Arctic warming contribution relative to the tropics (Pithan and Mauritsen 2014; Hahn et al. 2021). In response to reduced Arctic sea-ice in the PAMIP runs, surface albedo feedback activates in the summer season when Arctic surface warming and AA are weak and is inactive in winter when AA and Arctic warming are enhanced. This suggests that surface albedo feedback makes a negligible direct contribution to AA and that annual-mean warming contribution analyses may not accurately represent the mechanisms underlying AA, which occurs mainly in the cold season. However, the increased summer absorption of SW radiation may contribute to winter oceanic heat release, thereby indirectly contributing to AA as shown by Dai and Jenkins (2023). We also showed that bottom-heavy moistening related to Arctic sea-ice loss has negligible or slightly negative effects on the TOA radiative flux partly due to near-zero or negative values of the TOA LW water vapor kernel in the Arctic lower troposphere (Pendergrass et al. 2018). Future work may consider estimating water vapor feedback from the surface (rather than TOA) perspective to improve understanding of increased lower tropospheric moistening on the surface radiative flux.

We conclude that enhanced atmospheric energy convergence into the Arctic becomes an important process for weak Arctic warming and AA under global SST warming with fixed Arctic SIC. Arctic water vapor feedback also activates in response to remote SST warming rather than local SIC loss likely due to higher sensitivity of water vapor LW effects to upper-tropospheric than lower tropospheric moistening. Sea-ice loss and cold-season oceanic energy release are necessary to produce large bottom-heavy warming and AA, triggering Arctic positive lapse rate feedback. We also found a positive winter cloud feedback in response to sea-ice loss but weak Arctic cloud feedback in the SST warming runs, suggesting that changes in local Arctic surface conditions have a larger influence on Arctic cloud properties than remote processes. The surface albedo feedback activates in response to Arctic sea-ice loss under fixed global SST in the summer months but does not make a large direct contribution to AA due to weak summertime warming in the Arctic. The results from the piSST-piSIC run (i.e., TOTAL), where SIC and SST are changed simultaneously to their historical states, are similar to the sum of piSST-pdSIC, pdSST-piArcSIC, and pdSST-piAntSIC (i.e, SUM) relative to present-day conditions. This suggests that the sum of the separate effects of SST and SIC perturbations approximately equal the combined effects of SST and SIC changes on surface warming, climate feedbacks, and atmospheric energy transport.

We recognize that there are limitations associated with atmosphere-only model runs as the ocean is treated as a boundary condition. Ocean-atmosphere coupling and the oceanic component of the poleward energy transport have been shown to play important roles in the atmospheric response to sea-ice loss (Deser et al. 2015; Tomas et al. 2016). Thus, future work may compare our feedback calculations to the results from models with a full-depth dynamical ocean to account for ocean feedbacks. Nevertheless, our results help to untangle the influence of background global warming related to global SST changes or large Arctic warming related to sea-ice loss on Arctic climate feedbacks.

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Author Contributions

M. T. Jenkins performed the analysis for this study, made the figures, and wrote the first draft of the manuscript. A. Dai and C. Deser helped improve the study, the manuscript and the figures.

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Data Availability Statement

The PAMIP model output used in this study can be downloaded from https://esgf-node.llnl.gov/search/cmip6/.

Ethics Approval

Not applicable.

Consent for Publication

The authors agree to publish the paper in *Climate Dynamics*.

Competing Interests

The authors declare no competing interests.

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